Cycles of Doming and Eruption of the Miocene Kisingiri Volcano, Southwest Kenya[1]

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ABSTRACT

Volcanic cycles of doming and eruption of the Miocene Kisingiri volcano produced three sedimentary cycles recorded in the volcaniclastic strata of Rusinga and Mfangano Islands, Lake Victoria, Kenya. Each of the three cycles began with the deposition of cobble and boulder conglomerates shed from the volcanically domed Precambrian basement, followed by deposition of pyroclastic and volcaniclastic strata, representing nephelinite-carbonatite eruption of the Kisingiri volcano. Volcanogenic strata produced by the first two cycles (lower two-thirds of the Rusinga Group) are predominantly fine-grained tuffs and medium-grained volcaniclastic deposits, indicating alluvial deposition from a low-relief volcanic edifice. The third cycle is dominated by boulder debris flows, lava flows, and minor tuffaceous beds (upper part of the Rusinga Group and overlying Kisingiri Group). This last cycle records the formation of the high-relief Kisingiri stratocone, much of which is preserved in the dissected flanks of the volcano. The first two cycles are recorded in distal apron deposits but are not well preserved in the core of the volcano. Second-order sedimentary cycles, consisting of fining-upward sequences (5–10 m thick) of granular tuffaceous sandstones and conglomerates that fine into siltstones, stacked floodplain paleosols, and airfall tuff beds, dominate the strata of the first two cycles. These fining-upward sequences represent alluvial aggradation that accompanied and followed eruptive episodes that were much shorter in duration than the main cycles of doming and eruption.

Introduction

Geologic histories of volcanoes are commonly difficult to ascertain fully due to the complex and destructive nature of volcanic processes. Cores of volcanoes rarely preserve a complete record of volcanic events due to incomplete exposure and preservational bias. Distal deposits also record volcanic events; however, such records are commonly compromised by admixtures of material from different sources, by incomplete deposition, and by erosional events. In only a few cases has the synthesis of vent and distal volcanic facies been attempted (Swanson 1966; Hackett and Houghton 1989; Cas and Wright 1987). The Miocene Kisingiri volcano offers an excellent case study of the synthesis of vent and distal facies because it is an isolated, central vent volcano that sits on a peneplain of Precambrian granitic and metamorphic rocks, and because it is extensively dissected into a large amphitheater that exposes vent and flank facies and a large volcanic dome (figure 1). Arena and apron deposits of this volcano (*sensu* Pickford 1986) are exposed nearby on Rusinga and Mfangano Islands in Lake Victoria, where they have been extensively studied for their fossil anthropoids. In this paper, we synthesize the volcaniclastic deposits from the Rusinga and Kisingiri groups with the volcanic history deciphered from the exposed core and flanks of the Kisingiri volcano (McCall 1958; LeBas 1977). We interpret volcanic eruptive styles, cycles of doming and eruption, and alluvial response to pyroclastic volcanism through the identification and characterization of alluvial and pyroclastic deposits, granitic detritus, and paleosols in Rusinga Island strata.

Geologic Setting of Kisingiri Volcano

Large, central vent strato-volcanoes of nephelinite-carbonatite composition are common in East Africa and include Napak (King 1949; King 1965; Bishop 1968), Tinderet (Shackleton 1951; Deans

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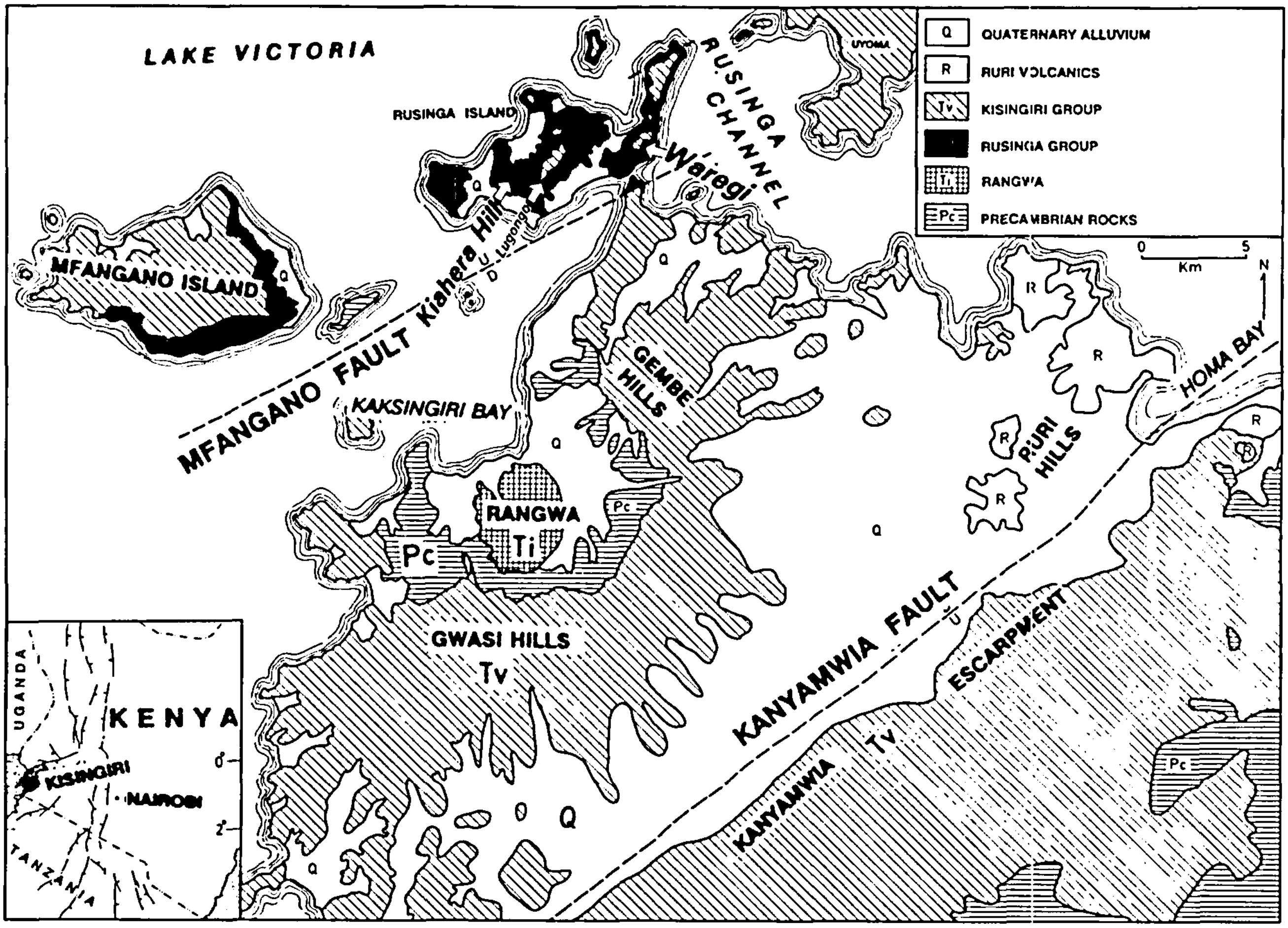

Figure 1. Location map of the Kisingiri volcano and Rusinga and Mfangano Islands and location of stratigraphic sections used for the composite section of figure 2.

and Roberts 1984), Shombole (Baker 1963; Peterson 1989), Kerimasi (Hay 1983; Mariano and Roedder 1983), and Oldoinyo Lengai, the only volcano in the world known to have erupted carbonatite tephra and lava in historic time (Dawson 1962; Dawson et al. 1987; Hay 1989). These volcanoes are associated with the East African Rift system and its associated alkaline volcanic province (Baker 1987; LeBas 1987). The East African Rift system was initiated during the early Miocene with broad epeirogenic uplift, eruption of flood lavas in the Turkana region, and the development of several nephelinite-carbonatite volcanic centers in eastern Uganda and western Kenya, including Kisingiri (Baker et al. 1971, 1972; Baker 1986). The dissected stratocone of Kisingiri has received much attention due to its well-exposed volcanic core and central dome (McCall 1958; Lippard 1973; LeBas 1977); such extensive study has made it one of the best known of the large nephelinite-carbonatite volcanoes.

The South Nyanza area bordering Lake Victoria in western Kenya is dominated by a regional slope underlain by the Precambrian Nyanza complex. This surface slopes westward from the rim of the Gregory Rift to the floor of Lake Victoria. The main mass of the Kisingiri volcano sits in a graben bounded on the northwest by the Mfangano fault and on the southeast by the Kaniamwia fault (figure 1). This graben is an extension of the Nyanza Rift (formerly Kavirondo Rift, Pickford 1982), which is an E-W rift lying west of the central part of the N-S Gregory Rift (figure 1). The Kisingiri volcano is now largely dissected but still retains volcanogenic features characteristic of large, central vent nephelinite-carbonatite volcanoes. A central domed area, consisting of the Rangwa intrusive complex and fenitized Precambrian rocks, is surrounded on three sides by an amphitheater-like valley and ridge (Shackleton 1951; McCall 1958; LeBas 1977). Precambrian rocks comprise the lower half of the ridge while outward-dipping

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strata of tuff breccia and lava flows comprise the upper portion of the ridge. The distal, fossil-rich volcaniclastic deposits of Rusinga and Mfangano Islands lie on the northwestern, up-thrown block of the graben rim (Shackleton 1951; VanCouvering and Miller 1969; Pickford 1986; Drake et al. 1988).

Cycles of Doming and Eruption

The Rusinga Group is the lower of two groups that resulted from the growth of the Kisingiri volcano (Shackleton 1951; Whitworth 1953; VanCouvering and Miller 1969; VanCouvering 1972; Pickford 1984; Pickford 1986; Drake et al. 1988). The Rusinga Group consists of fluvial and pyroclastic strata, while the overlying Kisingiri Group consists of debris flow deposits, nephelinite lava flows, and tuff breccia. Although the volcanic remnants of the Kisingiri volcano allow for the reconstruction of a large stratocone, estimated to have had a relief of between 2500 and 3000 m and a diameter of 70 km (LeBas 1977; VanCouvering 1972; Drake et al. 1988), the majority of the deposits in the Rusinga Group contain sedimentary features more indicative of a low-relief volcanic edifice. The change from the tuff- and sand-dominated Rusinga Group to the debris flow and lava flow-dominated Kisingiri Group represents progradation of proximal volcanic facies during construction of the Kisingiri stratovolcano (Pickford 1986; Thackray and Bestland 1988; Thackray 1989; Bestland 1990).

Within the overall upward-coarsening sequence on Rusinga and Mfangano islands are three distinct stratigraphic intervals that contain abundant and coarse-grained granitic and metamorphic detritus eroded from the Precambrian basement (Thackray and Bestland 1988; Thackray 1989; Bestland 1990). These intervals, recognized in the five formations on Rusinga Island, are interpreted to be part of sedimentary cycles. Each cycle consists of a lower section rich with Precambrian detritus and an upper section consisting of volcaniclastic and pyroclastic deposits (figure 2).

The composite section of Rusinga Island strata (figure 2) was constructed from the thickest stratigraphic sections on the island. The Wayondo Formation, Kiahera Formation, and Rusinga Agglomerate were measured at Kiahera Hill (Bestland 1990; Bestland and Retallack 1993). The lower part of the Hiwegi Formation was measured at Waregi (Bestland 1990; Retallack et al. 1995), and the Kibanga Member and the Kiangata Agglomerate were measured at Lugongo (Thackray 1989). Stratigraphic units were described in detail; field and lithologic characteristics such as tuffaceous component, percentage of basement-derived detritus, and degree of paleosol development were determined (Thackray 1989; Bestland 1990; Retallack et al. 1995). The tuffaceous component was evaluated on the content of fine-grained ashy matrix and the presence of accretionary lapilli (Reimer 1983), common in airfall beds of the Rusinga Group. Primary tuffs were recognized on the basis of graded to massive bedding, soft sediment deformation structures, and a lack of fluvial or paleosol features. The paleosol development index in figure 2 is based on field characteristics such as clay content, strength of clay structures, stage of calcareous nodule development, and color (Retallack 1988, 1990) and is substantiated, for selected paleosols throughout the section, with thin section point counting for grain size and with bulk-rock geochemical analysis (Bestland 1990; Thackray 1989; Bestland and Retallack 1993; Retallack et al. 1995).

Cycle I. Cycle I is recorded in the Wayondo and Kiahera Formations (figures 2, 3). The Wayondo Formation, lowermost of the Rusinga Group, consists of basement-derived conglomerate beds dominated by granitic and gneissic clasts. The overlying Kiahera Formation contains nephelinitic sandstones and siltstones and carbonatite-nephelinite airfall tuffs. The Wayondo Formation has long been interpreted as a product of the initial doming of Precambrian basement by intrusion of alkaline silicate magmas (Shackleton 1951; McCall 1958; LeBas 1977; Drake et al. 1988). These intrusions fenitized and brecciated the Precambrian granitic and metamorphic rocks (LeBas 1977). Intrusion of carbonatite and other alkaline rocks, coupled with eruption of pyroclastic material from small vents, followed the first phase of fenitization; the eruptions are the earliest pyroclastic activity recorded in the volcanic core (LeBas 1977). A volcanic hiatus at the end of Cycle I is recorded by a thick sequence of moderately developed paleosols at the top of the Kiahera Formation (pisolithic red earths of Shackleton 1951, remapped by Bestland and Retallack 1993). This series of non-tuffaceous paleosols correlates with a period of erosion of the Kisingiri volcano identified by LeBas (1977) during which intrusions and fenitized basement rock were exposed.

Cycle II. Cycle II is dramatically recorded in the Boulder Breccia Member, Rusinga Agglomerate, Kulu Formation, and the lower two-thirds of the Hiwegi Formation (figures 2, 3). Debris flow deposition of abundant and coarse granitic detritus in the Boulder Breccia Member (Drake et al. 1988, remapped by Bestland and Retallack 1993) begins the first stage of the second cycle. The 40 m thick Ru-

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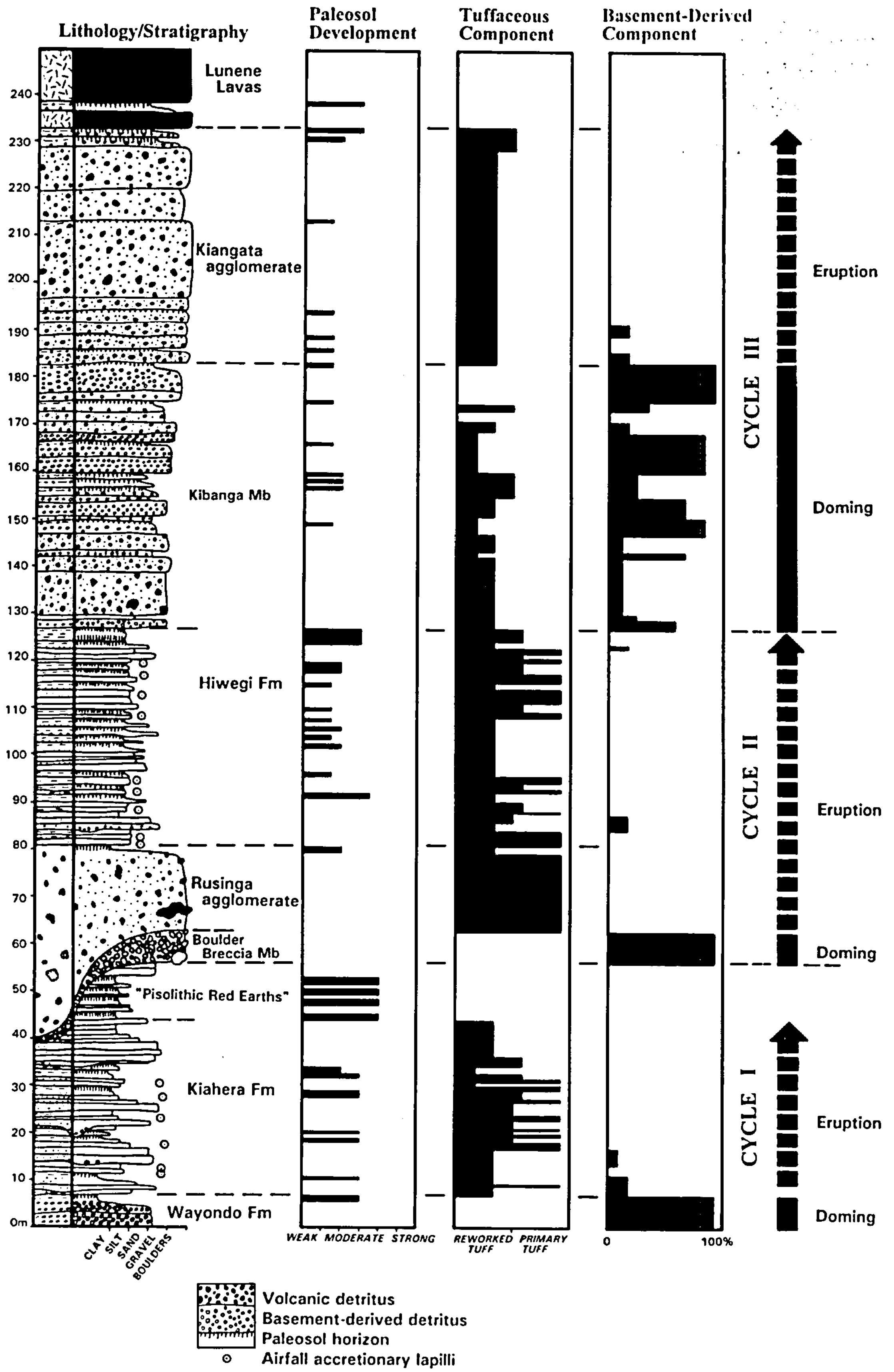

Figure 2. Composite stratigraphic section of the Rusinga and Kisingiri groups from Rusinga Island and corresponding paleosol development, pyroclastic input, granitic input, and interpretation of the three volcanic cycles.

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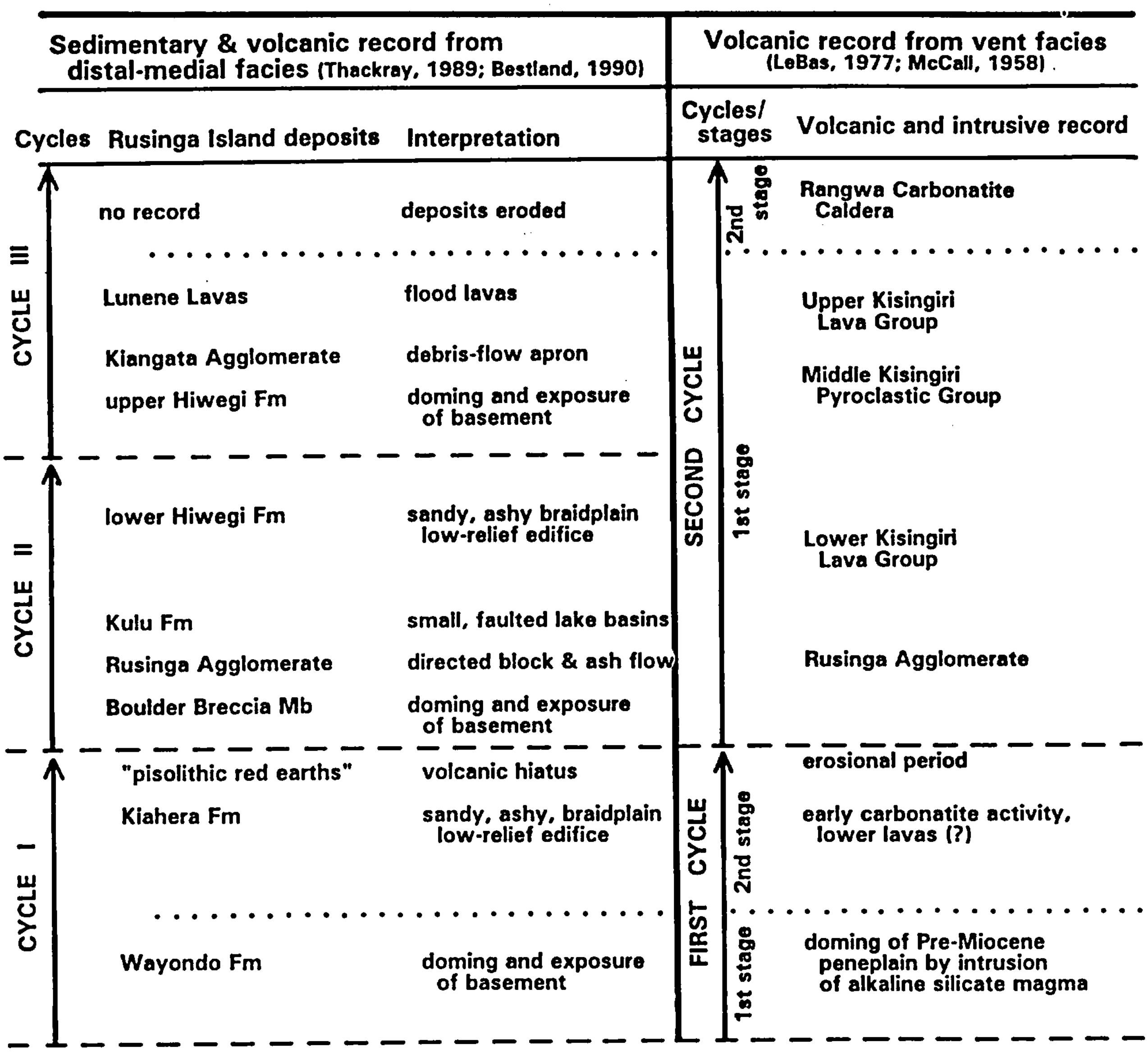

Figure 3. Comparison of the sedimentary and volcanic record from distal, volcaniclastic facies with the volcanic and intrusive record from Kisingiri vent facies.

singa Agglomerate tuff-breccia sheet, which overlies the Boulder Breccia Member, was produced by an "explosive eruption from a vent drilled through an ijolite complex" (LeBas 1977, p. 43). This eruption probably occurred on the northern side of the volcano because the Rusinga Agglomerate is present only on Mfangano and Rusinga Islands and at the mainland localities of Nyamarandi and Kibibura, north of Rangwa (LeBas 1977). A period of faulting followed the deposition of the Rusinga Agglomerate in the Rusinga Island area, as recorded by fan deltas of the Kulu Formation (Bestland 1991). Pyroclastic activity continued during this period (Bestland 1991) as did deposition of minor amounts of granitic detritus. Quasi-continuous pyroclastic and scoriaceous volcanism is recorded in the lower half of the Hiwegi Formation (figures 2, 3). The end of the second cycle occurs at the top of a series of paleosols that directly underlie thick, coarse, granitic conglomerates of the Kibanga Member (of VanCouvering 1972) of the Hiwegi Formation (Thackray 1989).

Cycle III. The onset of Cycle III is represented by the granitic clast-bearing Kibanga Member of the Hiwegi Formation (figures 2, 3). The overlying Kiangata Agglomerate and Lunene lavas cap the volcanic sequence preserved on Rusinga and Mfangano Islands. This sequence of coarse debris-flow

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deposits and nephelinite lava flows, with only minor tuffaceous interbeds and paleosols, records a distinctly different depositional regime than the Kiahera and Hiwegi formations. The coarse, clast-rich debris flows represent the apron facies, while the underlying sandy and tuffaceous strata of the first two cycles are part of the arena facies. The debris flow deposits of the Kiangata Agglomerate are a widespread unit around the Kisingiri volcano (McCall 1958) and record the growth of the Kisingiri volcano into a large, 3000 m stratocone. The overlying Lunene lavas are separated from the Kiangata agglomerate by a moderately developed reddish paleosol horizon. This paleosol represents a break in volcanism and supports the stratigraphic finding of McCall (1958) that a "strong erosional disconformity" exists between the mela-nephelinite flood lavas and the underlying nephelinite agglomerate exposed in the eroded core of the volcano.

Second Order Eruption Cycles. Smaller-scale sedimentary cycles are recognized in the Kiahera and Hiwegi formations. These second order cycles yield important information regarding eruption patterns during Cycles I and II. The second-order cycles consist of basal fine-grained calcareous tuffs with common accretionary lapilli and are directly overlain by granular tuffaceous sandstones and pebbly conglomerates that fine upward into siltstones, and finally stacked paleosols (figure 4). Airfall accretionary lapilli tuff beds are common in the stacked paleosols. Petrography and geochemistry of these calcareous tuffs indicate that they are melilitite tuffs with a minor carbonatite component, similar to melilitite-carbonatite tuffs described from the Pliocene-age Laetolil Beds by Hay (1978). The scoriaceous conglomeratic deposits associated with the tuffs in the basal part of the fining-upward sequences are nephelinitic with a component of melilitite but largely lack carbonatite tuff (Bestland 1990, 1991; Bestland and Retallack 1993).

These fining-upward sequences were recognized by Bestland (1990) and Bestland and Retallack (1993) but were previously used by Shackleton to subdivide the Kiahera Formation. The coarse-

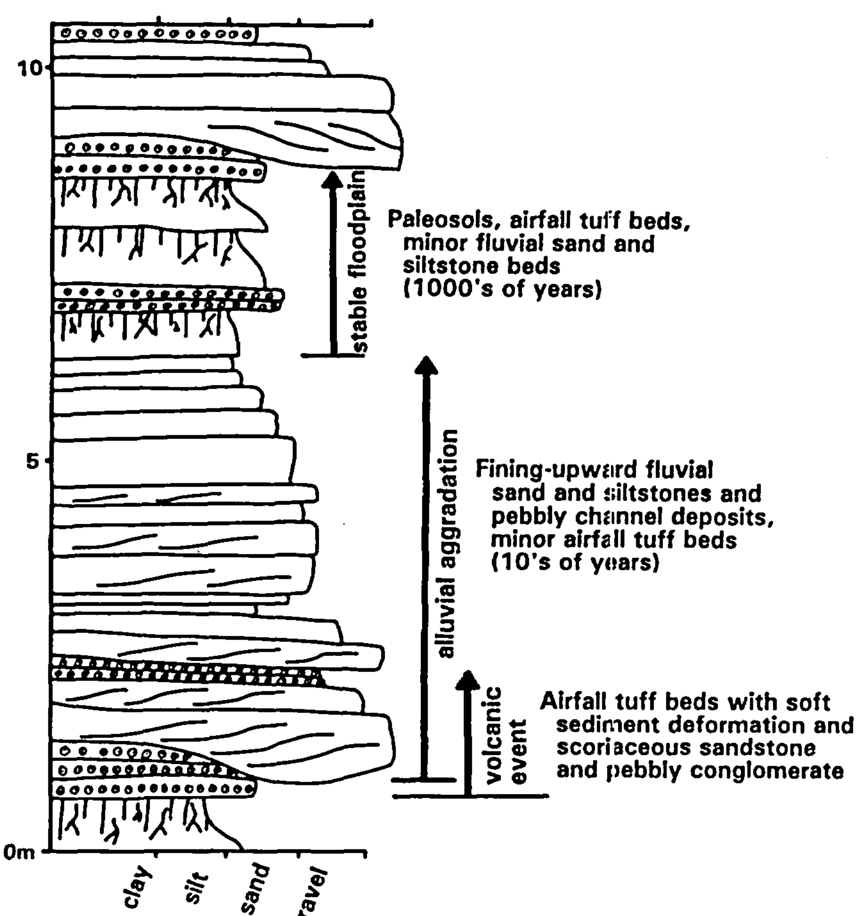

Figure 4. Composite stratigraphic section of second-order sedimentary cycles that make up the majority of strata in the Kiahera and Hiwegi Formations (Cycles I and II). At the base, the fining-upward sequences have channelized and incised pebbly conglomerates with braided stream gravel bar deposits with lateral accretion surfaces. These conglomerates commonly directly overlie calcareous airfall tuff beds. Sandy-gravel facies fine upward into overbank siltstone and fine sandstones which are capped by clayey and tuffaceous paleosols.

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grained basal parts of the fining-upward sequences were in part utilized by VanCouvering (1972) to subdivide the Hiwegi Formation. The basal channel bodies form ledges that crop out due to preferential erosion of clayey paleosol beds that underlie the more resistant pebbly sandstone beds. Low-angle cross-beds are common in the coarse sandstones and are interpreted as lateral accretion surfaces of sandy point bars (Bestland and Retallack 1993). The laterally continuous channel bodies are commonly overlain by levee-floodplain siltstones and fine sandstones. These overbank fine sandstones and siltstones are bedded and cross-bedded; they contain airfall tuff beds but do not contain evidence of soil formation. Another sedimentary type within the coarse-grained basal unit, although much less common than fluvial channel conglomerates, are vaguely bedded, graded pebbly sandstones and pebbly conglomerates interpreted as hyperconcentrated flood flow deposits (in the sense of Smith 1986; Nemec and Muszynski 1982). This fining-upward tuffaceous and scoriaceous package consequently represents rapid pyroclastic and alluvial aggradation that accompanied and followed volcanic eruption similar to other well known cases of volcanically influenced fluvial sedimentation (Vessel and Davies 1981; Smith 1987). The stacked paleosols sandwiched between these packages represent periods of volcanic hiatus when the floodplain was stabilized by vegetation, and sedimentation rates were lower.

Discussion

Two distinct stages of volcano growth and morphology are interpreted for the Kisingiri volcano based on the sedimentology of volcaniclastic deposits from the Rusinga Island section (figure 5). Deposits of Cycles I and II formed from volcanogenic material shed off a low-relief volcano, whereas deposits of Cycle III represent the growth of the high-relief Kisingiri stratocone. Volcaniclastic features of Cycles I and II deposits indicative of distal facies deposition include the abundance of fluvial sandstone and siltstone beds and fine-grained airfall tuffs, and the almost total lack of debris flow deposits. Debris flows are common on "ring plains" (*sensu* Hackett and Houghton 1989) or "arena" facies (Pickford 1986) associated with large stratocones where they fill and overflow gullies and channels cut into pre-existing deposits. Lahar deposits can be rapidly reworked by fluvial processes; however, coarse lag deposits remain (Waldron 1967), as do much of the overbank or floodplain facies of the debris flow (Scott 1988). Rusinga Island lies 15 to 20 km from the vent area, where debris flows would be expected given a moderate-sized strato-volcano. Distances of 15–20 km along the South Fork of the Toutle River during the 1980 Mt St Helens eruption contained high-energy, coarse debris flows, which in some cases abraded the channel (Scott 1988). Similar studies of debris-flow runout distances from Irazu Volcano (Waldron 1967), and Volcan Fuego (Vessel and Davies 1981) document similar high-energy debris-flows at these distances. These debris flows originated from volcanoes with 1500 m to 3000 m of relief. Following these lines of reasoning, the "low-relief volcanic edifice" of Kisingiri Cycles I and II was that of multiple scattered vents atop a domed Precambrian basement, all with relief of probably no more than 1000 m (figure 5).

The abundance of second-order eruption cycles in the deposits of Cycle I and II allows for the interpretation of a common eruption-sedimentation scenario. The stacked paleosols overlain by melilitite-nephelinite airfall beds record a period of volcanic hiatus followed by pyroclastic eruption. This pyroclastic eruption, or the continuation of the eruption, then caused alluvial aggradation of the ring plain by a rapidly migrating braided alluvial system, producing the lateral channel sandstones and pebbly conglomerates. As the eruption wanes, alluvium is reworked from proximal ring plain areas to lower ring plain settings, where it is deposited as the fining-upward siltstones and sandstone overbank deposits. With time, the channels are stabilized and deposition of floodplain levee facies occurs more slowly allowing for weak soil development. Fluvial incision of the ring plain during the following phase of volcanic hiatus isolates much of the ring plain and allows soils to develop. Vertical accretion of these ring plain soils by the addition of small increments of ash and dust buries the soils before they can develop mature profiles.

Also recorded in many of the second-order cycles is a lag time between the onset of volcanism and fluvial aggradation. This lag time is represented by the common occurrence of airfall tuff beds immediately underlying coarse channel sandstones (figure 4) and may be caused by the time required for the alluvial system to respond to pyroclastic and effusive volcanism. An ash-mantled landscape denuded of vegetation would suffer dramatically increased erosion rates and flood peaks and cause a pulse of ring plain sedimentation as has been documented at Irazu Volcano (Waldron 1967). An alternative interpretation is that the sedimentary sequence is caused by a change in the style of volcanism. An initial explosive and pyro-

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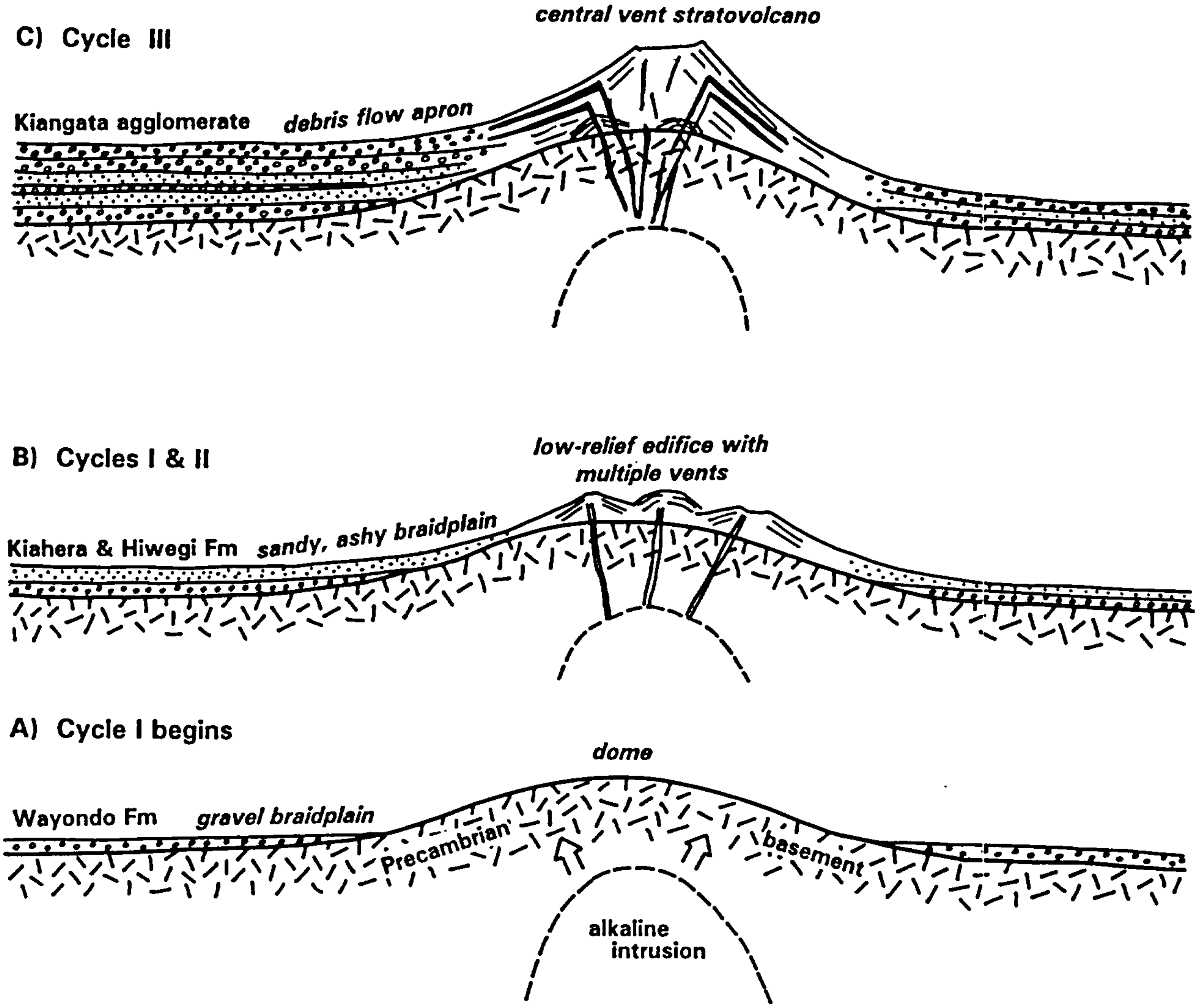

Figure 5. Reconstruction of volcanic and depositional events during the two-phase growth of the Kisingiri volcano. (*A*) Pre-volcanic doming of the Precambrian basement causes deposition of the conglomeratic and basement derived Wayondo Formation. (*B*) The first phase of volcanism (Cycles I and II) consists of scattered small vents atop the domed Precambrian basement. (*C*) The second phase of volcanism occurs during the growth of the Kisingiri stratovolcano associated with Cycle III.

clastic volcanic phase would produce the airfall beds and then would give way to mildly explosive and effusive volcanism represented by the channelized coarse tuffaceous sandstones.

Two cycles of eruption, each with two stages (figure 3), were recognized by LeBas (1977) from his study of the core of the Kisingiri volcano. These cycles began with the production of undersaturated alkaline silicate rocks and ended with a series of carbonatite pyroclastic and intrusive events (LeBas 1977). The three cycles recognized from the distal deposits (this paper) follow approximately the first cycle and half of the second cycle described by LeBas (1977). Each of the three cycles recognized in the distal facies is initiated with a doming phase, represented by basement-derived detritus, and presumably reflects the intrusion of alkaline silicate magma. Cycle II, recorded in the distal volcaniclastic deposits as the Boulder Breccia Member, Rusinga Agglomerate, Kulu Formation, and lower Hiwegi Formation, is not recognized in the core of the volcano by LeBas (1977). Lower Kisingiri Lavas occur beneath the Rusinga Agglomerate at scattered localities directly above the domed Precambrian basement (McCall 1958; LeBas 1977) and presumably reflect local accumulations of lavas around small vents. Lower Kisingiri Lavas also occur above the Rusinga Agglomerate; however,

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LeBas (1977, p. 95) states that "no distinction can be made between the lower lavas and the Middle Kisingiri Group." Thus, the Lower Kisingiri Lavas span a range of stratigraphic units including the Kiahera and Hiwegi Formations, and the Kiangata agglomerate.

The late-stage Rangwa caldera complex is interpreted to occur at the end of volcanic history of Kisingiri following mapping and interpretation by LeBas (1977). This 4 km diameter intrusive and caldera fill complex is not cut by any nephelinite dikes, as are other Kisingiri volcanics, and is a "single and complete" unit encased by basement rocks and early intrusive rocks (LeBas 1977, p. 101). This carbonatite caldera is apparently the last event recorded in the volcano and has not been identified in any distal deposits.

In summary, the synthesis of the distal volcanogenic deposits with the intrusive and effusive record preserved in the core of the volcano has produced a comprehensive history of the Kisingiri volcano. Two general phases of volcanic growth and morphology are deduced from the sedimentology of the Rusinga Island strata: a low-relief edifice with scattered vents that produced the fine-grained deposits of Cycles I and II, and a high-relief central-vent stratocone that produced the coarse-grained debris flows of Cycle III deposits. Large alkaline intrusions at depth domed the Precambrian basement, and produced evolved magmas and small scattered vents, responsible for the deposits of Cycles I and II. This magma body then centralized its vent during Cycle III to produce the Kisingiri stratovolcano and the Rangwa caldera.

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